Journal of Georgian Geophysical Society,
Issue (A), Physics of Solid Earth, v.15a, 2011-2012, pp. 88-97

Dynamic triggering of local seismic activity in Georgia by the great 2011 Japan earthquake

T. Chelidze, N. Zhukova, A. Sborshchikov, D. Tepnadze
M. Nodia Institute of Geophysics of Iv. Javakhishvili State University

Abstract

Introduction of new sensitive broadband seismographs, new dense seismic networks and new methods of signal processing lead to the breakthrough in triggering and synchronization studies and formation of a new important domain of earthquake seismology, related to dynamic triggering of local seismicity by wave trains from remote strong earthquakes. In the paper are considered the peculiarities of triggered seismicity in Georgia on the example of 11.03.2011great Tohoku earthquake in Japan. (M=9,) and moderated earthquake in East Greece (09. 03.2011).

The study of seismic response of the lithosphere to a weak forcing is a fundamental problem for seismic source theory as it reveals the important detail of the tectonic system, namely, how close is it to the critical state. Last years introduction of new sensitive broadband seismographs, new dense seismic networks and new methods of signal processing lead to the breakthrough in triggering and synchronization studies and formation of a new important domain of earthquake seismology, related to dynamic triggering (DT) of local seismicity by wave trains from remote strong earthquakes (Hill, Prejean, 2009; Prejean, Hill, 2009; Hill, 2010). The trivial aftershocks' area is delineated mainly by static stress generated by earthquake and decay rapidly with distance d as d^{-3}, whereas the dynamically triggered stresses decay much slower (as $d^{-1.5}$ for surface waves). That means that dynamic stresses generated by seismic wave trains can induce local seismicity quite far from the epicenter; they can be defined as remote aftershocks. The first well documented DT episode is connected with 1992 Landers earthquake, when the sudden increase of seismicity above background value (calculated as β-statistic of Matthews and Reasenberg, 1988) after the main event was observed by many seismic stations at distances up to 1250 km with delays ranged from seconds to days. Later on DT was observed in different remote areas after Denali Fault 2002, Hector Mine 1999, Kurile 2007, Sumatra, 2004 and many other EQ, though most clearly the effect is expressed in active extensional regime areas, as well as in volcanic and geothermal regions.

The main characteristic of DT events are peak dynamic values of stress (T_p) or strain (ε_p); for shear waves $T_p \approx G\ (u_p/v_s)$ and $\varepsilon_p \approx u_p/v_s$; here G is the shear modulus, u_p is particle' peak velocity and v_s is velocity of the shear wave. Calculated from the field data give values of T_p from 0.01MPa to 1MPa (ε_p from 0.03 to 3 microstrain). Such large scatter is due to the impact of another important factor, namely, the local (site) strength of earth material, which is highly heterogeneous. Thus what matters is not the absolute value of T_p or ε_p, but the difference between local stress and local strength (resistance to failure). This is why in some areas high T_p do not trigger local seismicity and, on contrary, some areas manifest DT even at low peak stresses. One of main factors reducing local strength is the pore pressure of fluids, which is the scope of relatively new direction, so called hydroseismology (Costain and Bollinger, 2010).

The stresses imparted by teleseismic wave trains according to assessments of D. Hill (2008) are 10^5 times smaller than confining stresses at the depth, where the tremors are generated. This is not surprising as the synchronization theory predicts that even smallest forcing is able to adjust the rhythms of oscillating systems (Pikovsky et al, 2003)

In most cases triggering is observed during surface waves, especially during Rayleigh wave arrivals, i.e. long periods and large intensity of shacking are favorable for exciting remote triggered events. Periods in the range 20-30 sec are considered as most effective in producing triggered events for the same wave amplitude. In principle the optimal period of DT should depend on the earthquake preparation characteristic time and can change from dozens of seconds for microearthquakes to hours and days for moderate events. For tidal stresses with periods 12-24 h the threshold can be as low as 0.001 MPa.

Timing of triggered events is very variable: they can be excited immediately by the some phase of the wave train (say, Rayleigh) or delayed by quite a long time, hours or days. Duration of triggered activity period is also variable - from minutes to a month.

Magnitude of reported triggered events varies between M = 0.2 or less to M = 5.6. It is likely that most of triggered seismicity are just ignored due to their small intensity and are not included in seismic catalogs. Small (local) triggered events in a given area are revealed using very simple method: the original record of the strong (remote) earthquake are filtered in order to separate low-frequency component (0.01-1 Hz), i.e the dominant component of passing wave train, which can be considered as a forcing and high-frequency component (1-20 Hz), where local triggered events can be recognized.

The triggered events belong to one of two classes: regular earthquakes with sudden onset and so called non-volcanic tremors or tectonic tremors (TT) with emergent onset.

Tectonic tremors are considered as a new class of seismic events related to recently discovered phenomena of low frequency earthquakes and very low frequency earthquakes (Obara, 2003). As a rule individual tremor has dominant frequencies in the range 1-10 Hz, lasts for tens of minutes and propagates with shear wave velocity, which means that they are composed by S body waves. Spatially triggering is most frequently encountered in hydrothermal areas

At present a lot of instances of triggering and synchronization are documented using statistical approach, but the most informative technique is the above mentioned double-filtering method. As a rule, triggered events belong to the class of triggered tremors. Tremor's signatures are: emergent onset, lack of energy at frequencies higher than 10 Hz, long duration from dozens of seconds to several days, irregular time history of oscillations' amplitude, close correlation with large-amplitude surface waves.

Of course, different patterns can be observed also. For example great Tohoku M= 9 earthquake, Japan, triggered local seismic events (Figs. 1 a, b) in Georgia (Caucasus), which is continental collision area, separated from Japan by 7800 km. Recorded seismic waves were converted to WAV format with the corresponing sampling rate using tools provided in MATLAB application.

As the Caucasus is dominated by compression tectonics and the triggering examples from such areas are rare, presented data are significant for understanding trigger mechanisms. High pass (0.5-20 Hz) filtered records at two broadband seismic stations located in Oni (South slope of Greater Caucasus) and Tbilisi (valley of river Kura), separated by the distance 130 km show that in this case the strongest triggered event at both sites corresponds to arrival of *p*-wave instead of surface waves. The sequence of triggered events is quite similar at both stations. Tbilisi is a hydrothermal area and so it falls into general class of triggering-prone regions, but Oni is not a hydrothermal area. Here the fracture can be promoted just by pore fluid pressure.

The comparison of three components of records (N, E and Z) shows that (Fig.2, 3): i. on horizontal components (E and H) triggered events, besides p-arrival are also generated by Love and more intensively by Rayleigh waves; ii. vertical component (Z) generates tremors only at arrival of *p*- and Rayleigh waves, as it could be expected.

introduction of new sensitive broadband seismographs, new dense seismic networks and new methods of signal processing lead to the breakthrough in triggering and synchronization studies and formation of a new important domain of earthquake seismology, related to dynamic triggering (DT) of local seismicity by wave trains from remote strong earthquakes.

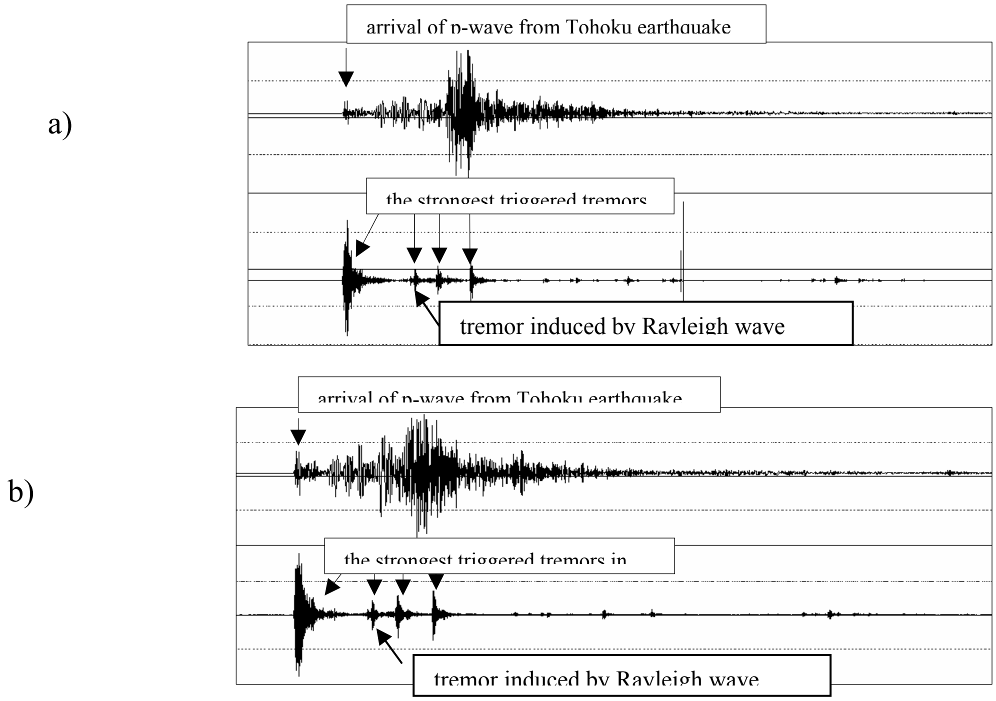


Fig. 1. Broadband record of M= 9 Tohoku EQ, Japan (11.03.2011) wave train z-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel). Arrows mark p-wave arrival. The lower channel shows local triggered events; the strongest event corresponds to arrival of p-wave. a. Oni and b. Tbilisi seismic station.

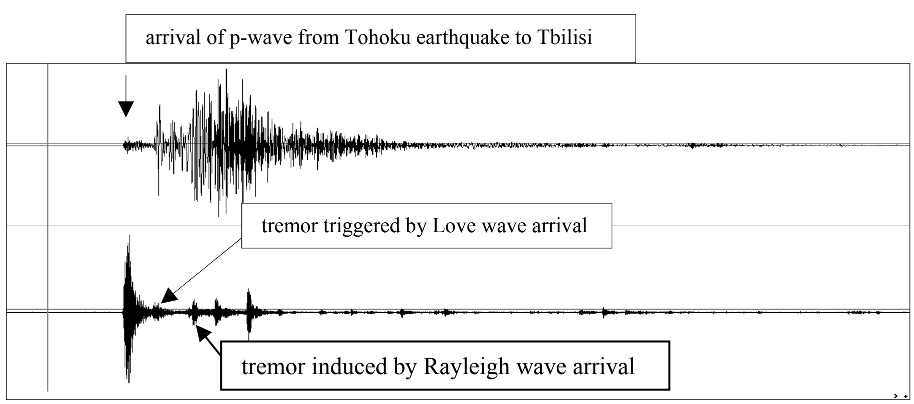


Fig. 2. Broadband record of M= 9 Tohoku EQ, Japan (11.03.2011) wave train N-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel) in Tbilisi. Arrows mark p-wave arrival. The lower channel shows local triggered events; the strongest event corresponds to arrival of p-wave. Here the Love wave also generates relatively weak tremor.

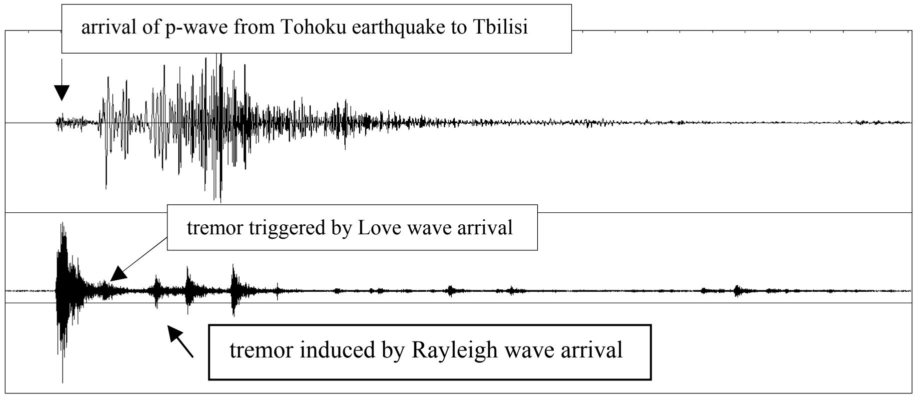


Fig. 3. Broadband record of M= 9 Tohoku EQ, Japan (11.03.2011) wave train E-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel) in Tbilisi. Arrows mark *p*-wave arrival. The lower channel shows local triggered events; the strongest event corresponds to arrival of *p*-wave. Here the Love wave also generates relatively weak tremor.

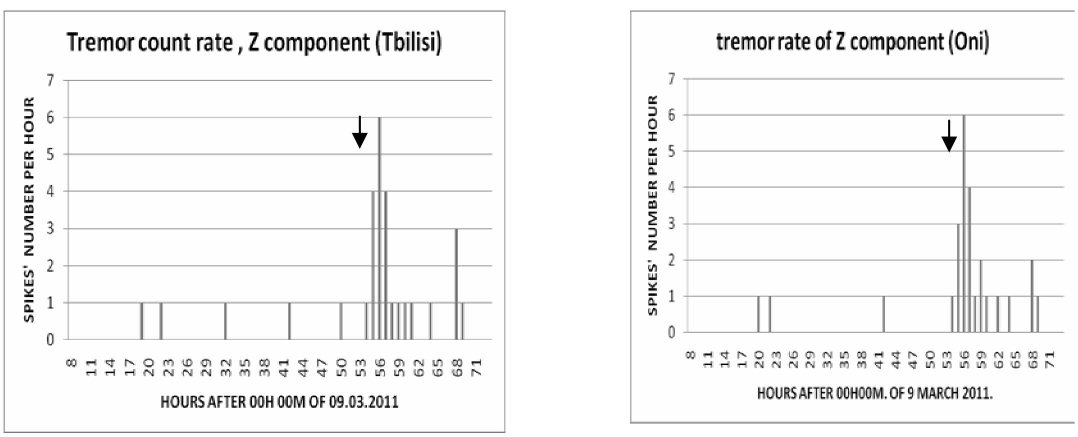
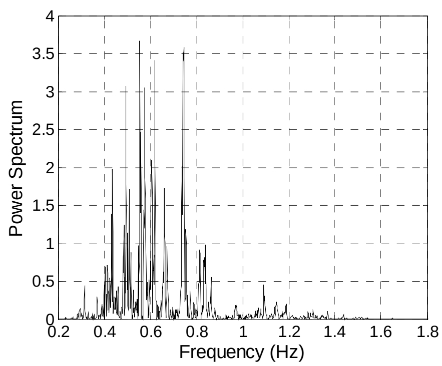


Fig. 4. Tremor rate (number of local events per hour) before, during and after Tohoku event in Tbilisi and Oni. Tohoku earthquake arrival time is marked by arrow.

The counting of tremors' rate (number of local events per hour) before, during and after Tohoku event both in Oni and Tbilisi reveals clear maximum just during the strong earthquake wave train passage, including coda (Fig. 4 a, b), which confirms the reality of triggering phenomenon. The duration of anomalously high tremor rate is of order of 6-8 hours.

Power spectrum of the triggered tremors shows that the maximal energy is released in the frequency range 0.4-0.8 Hz, i.e. these event are deficient at relatively high frequencies (Fig. 5 a). Tremor spectrum differs very much from the power spectrum of the broadband recording of Tohoku earthquake, which indicates that maximal power in Georgia was relieved at much lower frequencies, in the range 0.01-0.1 Hz (Fig.5 b). That means that very low-frequency forcing is necessary for triggering tremors. In other words, forcing of a period 100-10 sec is the time, necessary for tremor area activation.

It is interesting that not only strong earthquakes, but also middle size remote events also can trigger local earthquakes. For example, M=4.6 earthquake in East Greece (09. 03.2003) also triggered local seismicity in Georgia, separated from the epicenter by 1700 km, here again the strongest triggered event coincides with p-wave arrival (Fig. 6 a, b).



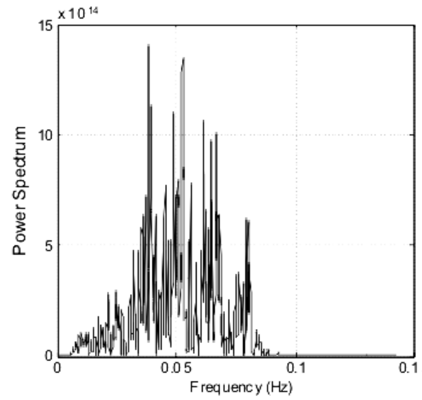
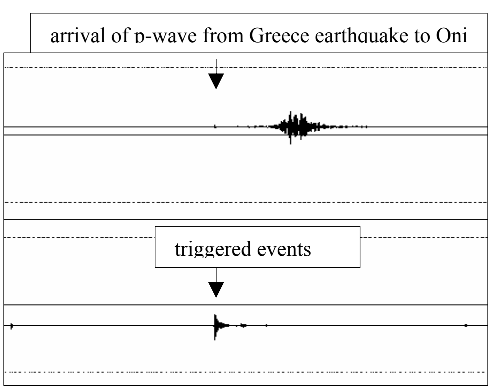
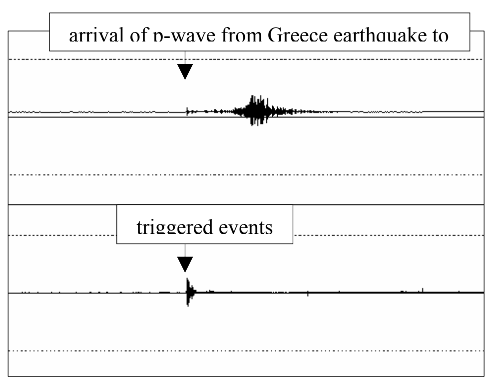


Fig. 5 a. Spectrum of the largest (first) triggered tremor in Tbilisi. Bandpass Butterworth filter was used to filter data in a range 0.5-20 Hz. b. spectrum of the broadband recording of Tohoku earthquake in Oni.





a. Oni b. Tbilisi

Fig. 6 a, b. Broadband record of M=4.6 earthquake in East Greece (09. 03.2011) wave train z-component (upper channel) and the same high-pass band (0.5-20 Hz) filtered record (lower channel).

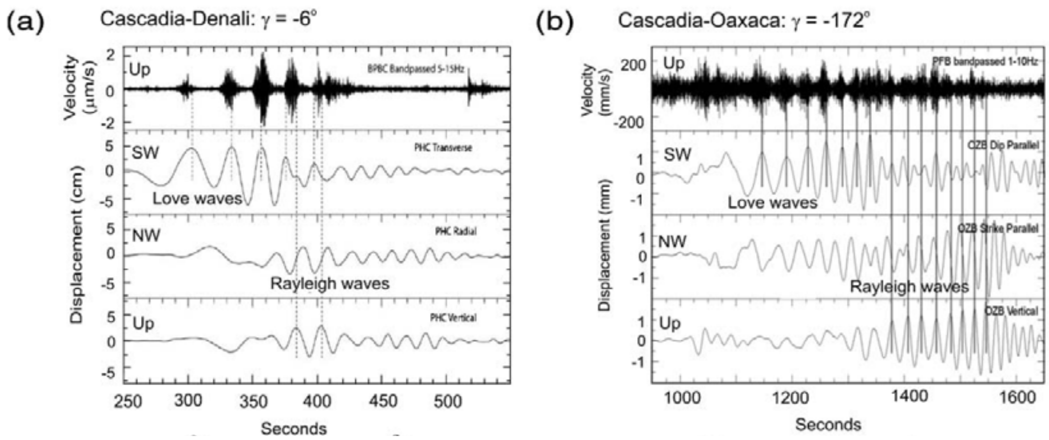


Fig.7. Examples of tremor triggered on the Cascadia megathrust beneath Vancouver Island, B.C., by surfaces waves from four Mw >7. 5 earthquake with incidence angles γ (Rubinstein et al. 2009; Hill, 2010). The top panel in each example shows broadband displacement waveforms for the incident surface waves (bottom three traces) and the high-frequency (5 to 15 Hz) traces for the triggered tremor (upper trace). (a) The Mw 7.9 Denali fault earthquake of 2002, tremor depth 15 km; (b) the Mw 7.5 Oaxaca earthquake of 1999, tremor depth 35 km

The lower channel shows local triggered events; the strongest event corresponds to arrival of p-wave. a. Oni and b. Tbilisi seismic station

Rubinstein et al. (2009) and Hill (2010) show clearly (Fig. 7a,b) that the weak forcing by wave train of remote strong earthquake can not only trigger, but also induce phase synchronization of induced events with surface waves.

The strong resemblance between our experimental results on electromagnetic (Fig. 8) or mechanical synchronization of stick-slip (Chelidze et al, 2006, 2007, 2010) and large scale natural events (Fig. 7) show that the phenomenon of synchronization has universal character and it can be successfully modeled in laboratory.

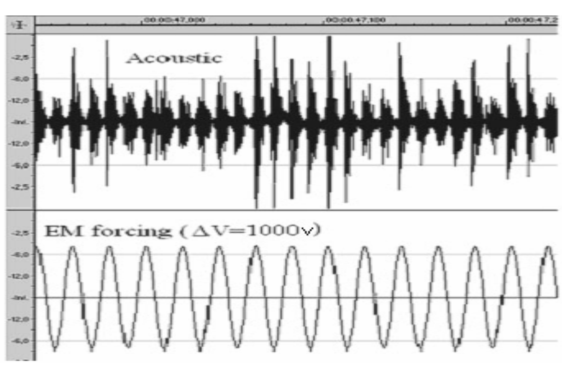


Fig.8. Acoustic emission (upper channel) during slip after application of 1000 V external periodical voltage (lower channel). Note complete phase synchronization between EM forcing and AE .

he physical mechanism of remote triggering is not clear. The mechanism should be different for triggered events closely correlated to wave train phase (direct triggering) and for delayed response.

Hill (2010) assessed (direct) triggering potential of wave trains from the fracture mechanics point of view, using Mohr and Coulomb-Griffits failure criteria. In general, Love waves incident on vertical strike-slip faults have a greater potential than Rayleigh waves, but the potential of Rayleigh waves incident on dip-slip faults dominates over Love wave potential. At the same time, the fault geometry and frictional strength are variable. Such heterogeneity leads to deviations from the above simple rule.

For large delays frictional failure, subcritical crack growth and excitation of crustal fluids are suggested as appropriate models (Hill, Prejean, 2009; Prejean, Hill, 2009; Hill, 2010).

We can stress close resemblance of our laboratory stick-slip experiments with typical recordings of ETS (Fig. 9); it seems that different morphology of the ETS signals can be explained by the various conditions of frictional motion, in particular, by different stiffness of dynamical system.

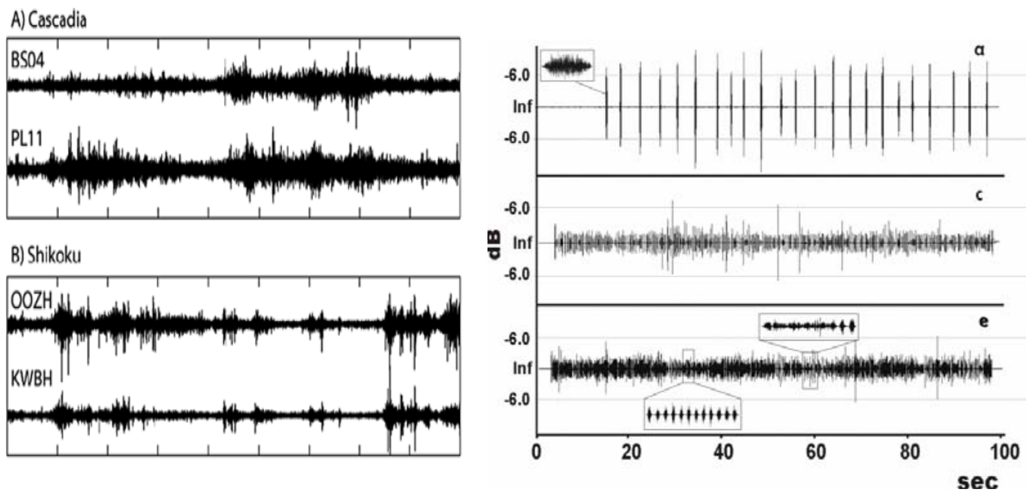


Fig. 9. Left side: recordings of non-volcanic tremor in the Cascadia subduction zone and the Nankai Trough. Records are bandpass filtered at 1–8 Hz. Right side: typical examples of AE recordings at different values of dragging spring stiffness: a) K= 78.4 N/m, c) K= 1068 N/m, e) K= 2000 N/m , f) K= 2371.6 N/m. Insets show AE wave train on extended time scales.

Obara (2002) and Rubinstein et al (2010) note that periods of tremor activity turn on and off by local or teleseismic earthquakes and remark that ‘no satisfactory model has been proposed to explain how teleseismic event might stop a period of active tremor”.

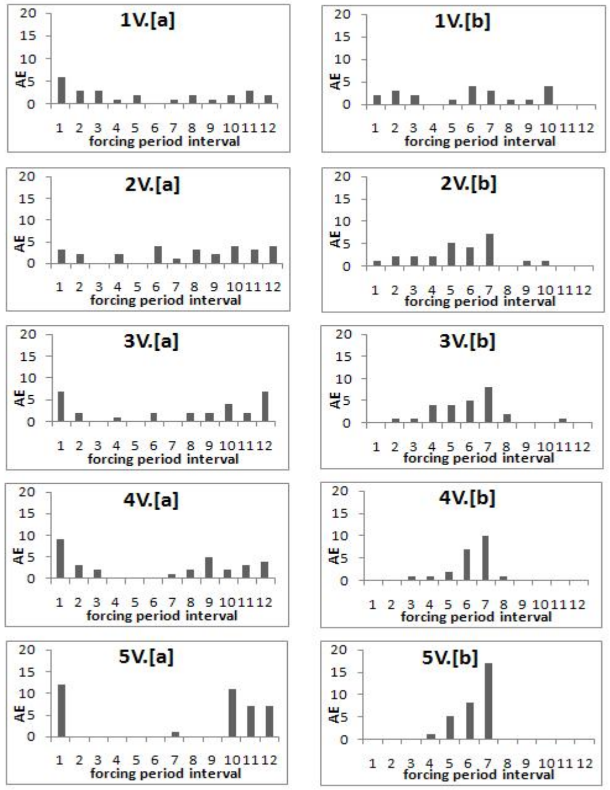


Fig. 10. Distribution of acoustic emission onsets (the left column) and terminations (the right column) relative to the (mechanical) forcing period phase (in twelfths of the forcing period) for different intensities of tangential forcing. Forcing frequency – 80 Hz.

The general explanation of how small-amplitude teleseismic wave can start of stop a period of tremor activity is the extremely high sensitivity of nonlinear systems to a weak forcing. The physical (laboratory) model of mentioned tremor arrest effect has been realized in our experiments with mechanical synchronization of stick-slip (Fig. 10). This remarkable result shows that very small mechanical forcing, 10^5 times smaller than the main driving force can affect both onsets and terminations of stick-slip generated acoustic wave train.

It seems that further development of sensitive devices, dense networks and processing methods will develop a new avenue in seismology, which can be defined as microseismology and which will study systematically small earthquakes and tremors, especially triggered and synchronized events. These events at present are ignored by routine seismological processing and are not included in traditional catalogues. At the same time, microseismic events contain very important information on geodynamics of processes and can give clues to understanding fine mechanism of nonlinear seismic process and may be, even contribute to the problem of earthquake prediction. Microseismicity can be compared by its importance to studies of elementary particles in physics.

References:

[1] Chelidze, T., O. Lursmanashvili, T. Matcharashvili and M. Devidze. 2006. Triggering and synchronization of stick slip: waiting times and frequency-energy distribution Tectonophysics, 424, 139-155

[2] Chelidze T., and T. Matcharashvili. 2007. Complexity of seismic process, measuring and applications – A review, Tectonophysics, 431, 49-61.

[3] Chelidze, T., Matcharashvili, T., Lursmanashvili, O., Varamashvili N., Zhukova, N., Meparidze. E. 2010. Triggering and Synchronization of Stick-Slip: Experiments on Spring-Slider System. in: Geoplanet: Earth and Planetary Sciences, Volume 1, 2010, DOI: 10.1007/978-3-642-12300-9; Synchronization and Triggering: from Fracture to Earthquake Processes. Eds.V.de Rubeis, Z. Czechowski and R. Teisseyre, pp.123-164

[4] Hill, D. Surface wave potential for triggering tectonic (nonvolcanic) tremor. 2010. Bull. Seismol. Soc. Am. 100, 1859-1878.

[5] Hill, D., Prejean, S. 2009. Dynamic triggering. In: Earthquake seismology, Volume editor H. Kanamori. Elsevier. pp. 257-293.

[6] Matthews, M. and Reasenberg, P. 1988.Statistical methods for investigating quiescence and other temporal seismicity patterns. Pure and Appl. Geophys. 126, 357-372.

[7] Obara, K. 2003. Time sequence of deep low-frequency tremors in the Southwest Japan Subduction Zone. Chigaku Zasshi (J. Geogr.)112, 837-849.

[8] Pikovsky, A., Rosenblum, M.G., Kurths. J. 2003. Synchronization: Universal Concept in Nonlinear Science. Cambridge University Press, Cambridge

[9] Prejean S., Hill, D. 2009. Dynamic triggering of earthquakes. In: Encyclopedia of Complexity and Systems Science, R. A. Meyers (Ed.), Springer, pp. 2600-2621.

[10] Rubinshtein et al. 2010, Non-volcanic tremors. In “New Frontiers in Integrated Solid Earth Sciences. S. Cloetingh, J. Negendank,(Eds), Spronger, Berlin, doi 10.,1007/1007/978-90-481-2737-5. pp. 287-314.

(Received in final form 20 December 2012)

Динамическое триггерирование локальных землетрясений в Грузии сильнейшим землетрясением 2011 года в Японии

Т. Челидзе, Н. Жукова, А. Сборщиков, Д. Тепнадзе

Резюме

Использование новых чувствительных широкополосных сейсмографов, современной плотной сейсмической сети, современных методов обработки сигналов привело к прорыву в изучении таких явлений как триггерирование и синхронизация, и формированию новой важной области в сейсмологии землетрясений, связанной с динамическим триггерированием локальных землетрясений серией волновых пакетов, приходящих от удаленных землетрясений. В данной статье рассмотрены примеры триггерируемой сейсмичности в Грузии на примере сильного землетрясения в Тохоку, Япония (11.03.2011, M=9) и среднего землетрясения в Восточной Греции (09.03.2011).

საქართველოში ლოკალური სეისმურობის დინამიკური ტრიგერირება დიდი 2011 წლის იაპონიის მიწისძვრით

თ. ჭელიძე, ნ. ჟუკოვა, ა. სბორშჩიკოვი, დ. ტეფნაძე

რეზიუმე

ახალი მაღალი გრძნობიარობის ფართოსიხშირული სეისმოგრაფებით აღჭურვილ;ი მჭიდრო სეისმური ქსელების შექმნამ და სიგნალის დამუშავების ახალი მეთოდების შემოტანამ განაპირობა გარღვევა მიწისძრების ტრიგერირების და სინქრონიზაციის კვლევაში. ფაქტობრივად შეიქმნა სეისმოლოგიის ახალი დარგი: ლოკალური სეისმურობის დინამიკური ტრიგერირება შორეული მიწისძვრების ტალღური პაკეტებით. სტატიაში განიხილება საქართველოში ტრიგერირებული სეისმურობის თავისებურობანი 11.03.2011 წლის დიდი ტოხოკუს (იაპონია) და 09.03.2011 აღმოსავეთ საბერძნეთის საშუალო სიძლიერის მიწისძვრების მაგალითზე.